Clouds and Radiation: A Primier

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This paper addresses a previously unknown complex interdisciplinary process providing a feedback loop which may have a major impact on the effect on global climate of the steadily increasing growth of greenhouse gases in the atmosphere.

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1 INTRODUCTION

Inclusion of the effects of clouds in climate models to a remarkably high degree of precision is almost certain to be a prerequisite to any reasonable degree of reliability in climate prediction because of the sensitivity of, for example, surface temperatures to cloud albedo. Unfortunately, many of the major physical processes taking place in clouds are still poorly understood. Some of them involve not only physics, but marine biology, oceanic and atmospheric chemistry, and other disciplines outside of physics, as well. It is likely, therefore, that not only do physical processes that are understood need to be modeled much more accurately than is now done, but many presently ignored processes have to be included as well. Indeed, this is the motivation behind DOE's Atmospheric Radiation Measurement (ARM) program.

This paper addresses, as an illustrative example, one such previously unknown complex interdisciplinary process providing a feedback loop which may have a major impact on the effect on global climate of the steadily increasing growth of greenhouse gases in the atmosphere. It is to be stressed that this is only one of a number of such feedback loops, many of which have probably not even been thought of yet, but all of which are entirely ignored in present day computer models.

Clouds influence climate in two major ways: they reflect incident short wavelength solar radiation thus preventing it from reaching the earth's surface and heating it, and they absorb outgoing long wavelength radiation from

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the earth thus reducing the earth's ability to cool itself. Which of these two competing, opposite sign, effects dominates is a sensitive function of the interaction of clouds with radiation, which is itself a sensitive function of the processes going on within various kinds of clouds.

Solar radiation is (to a good approximation) a black-body spectrum at $5{,}780°K$, which peaks in the visible at a wavelength of about 0.5 μm. At the earth, the flux of solar radiation is 1,370 watts/m^2, and due to the existence of day and night and because the earth is a sphere, the average solar energy incident at the top of the atmosphere is one fourth of this. The average earth albedo is .3, so 30% of the incident radiation is reflected, leaving 240 watts m^2 to be absorbed by the earth's surface and atmosphere.

Radiation emitted from the earth is (to a relatively poor approximation) a black-body spectrum at $255°K$, which peaks in the infrared at about 10 μm. There is essentially no overlap between the solar spectrum and the earth's spectrum. As the outgoing IR radiation passes through the atmosphere, some of it is absorbed and reradiated back toward the earth, but of course the net IR radiation escaping into space from the top of the atmosphere is also 240 watts/m^2, since the earth is (essentially) in equilibrium.

The black-body spectra and the principal atmospheric absorption bands are shown in Figure 1.

The albedo of the earth's surface (insofar as it can be measured from satellites) is no more than .15. The average cloud cover over the earth is observed to be 60% (65% over the ocean and 50% over the land). Hence, to

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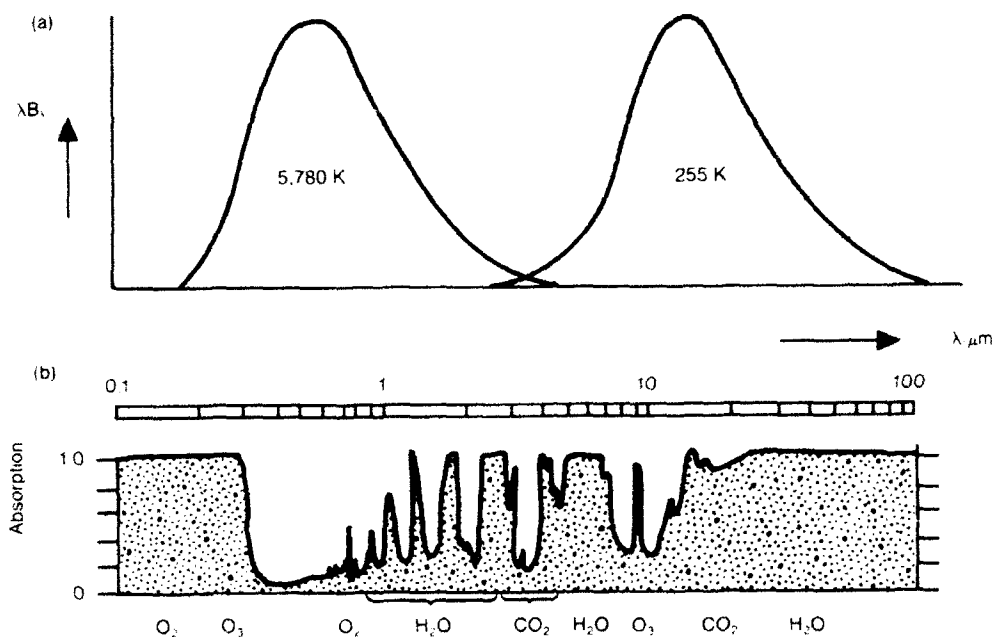
Figure 1. Curves of black-body energy B_λ at wavelength λ for 5,780 K (approximating to the sun's temperature) and 255 K (approximating to the atmosphere's mean temperature). The curves have been drawn of equal areas since integrated over the earth's surface and all angles the solar and terrestrial fluxes are equal. Absorption by atmospheric gases for a clear vertical column of atmosphere. The positions of the absorption bands of the main constituents are marked. (From R. M. Goudy "Atmospheric Radiation," Oxford, 1964.)

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make up the average albedo of .3, the albedo of clouds is about .45. Clouds are therefore the most important component in the amount of reflected sunlight, and their existence is crucial in determining the surface temperature of the earth. In fact, (if other parameters were held constant) a change of cloud albedo by 2% would warm the earth's surface by 1°C[1]. Evidently, the modeling of clouds in climate models must be done very accurately.

Clouds are composed of water droplets and/or ice crystals, which form on cloud condensation nuclei (CCN). (In principle, condensation of water vapor can occur at humidities above 100%, but because the vapor pressure increases with the curvature of the surface at which condensation occurs, in practice humidities of above 300% are needed for pure water vapor to condense. Therefore actual cloud formation takes place because of the presence of relatively large aerosols. These aerosols constitute CCN.) Clouds occur in several different forms, depending on their altitude, formation mechanisms, hydrodynamic properties, etc. Table 1[2] shows the distribution of the various forms over land and ocean. The relative amounts of these types often have diurnal variations which are of considerable importance. For example, the ratio of stratus to stratocumulus varies from less than 20% at local noon to nearly 30% in the early morning, while cumulonimbus peaks at 30% in late afternoon or evening and vanishes early in the morning[2].

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Table 1

Average % of Sky Cover over { Ocean / Land
of Various Cloud Types

Cumulus	12
	5
Cumulonimbus	6
	4
Stratus and Stratocumulus	34
	18
Nimbostratus	6
	6
Altostratus and Altocumulus	22
	21
Cirrus	13
	23

2 INCIDENT SOLAR RADIATION

For incident solar radiation, emission is irrelevant because at $\lambda \sim 0.5\mu$m. the black body spectrum at the earth's temperature of around 270°K is essentially zero. So only scattering and absorption count. To a first approximation, let us neglect scattering into the beam, and assume the incident solar radiation suffers only losses due to clouds. Thus the radiation intensity $I_\lambda(\hat{k}, z)$ of photons of wavelength λ travelling in the direction $\hat{k}$ satisfies[3]

$$dI_\lambda(\hat{k}, z) = -(\sigma_\lambda^{\text{abs}}(z) + \sigma_\lambda^{\text{scatt}}(z))N(z)I_\lambda(\hat{k}, z)dz, \qquad (2-1)$$

where σ are absorption and scattering cross sections and N is particle density. Therefore,

$$I_\lambda(\hat{k}, z_2) = I_\lambda(\hat{k}, z_1)e^{\tau(z_2, z_1)}, \qquad (2-2)$$

where the (dimensionless) optical depth τ is

$$\tau(z_2, z) = \int_{z_1}^{z_2} \sigma_\lambda^e(z)N(z)dz \qquad (2-3)$$

and

$$\sigma^e \equiv \sigma^{\text{abs}} + \sigma^{\text{scatt}} \qquad (2-4)$$

is the extinction cross section.

Each type of cloud has, at a given height z_1, a distribution $n(r, z)$ of sizes with particles of radius r (ice crystals are of course not spherically symmetric, though their absorption cross section does not differ markedly from that of liquid water droplets). In general, the cross section σ will also depend on the particle size. For wavelengths that are small compared to the particle size,

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as is typically the case for solar radiation (see Figure 2), the cross section is twice geometrical: $\sigma = 2\pi r^2$. Thus (if we take z_1 to be the top of the atmosphere where $\tau = 0$) the optical depth at height z is

$$\tau(z) = 2\pi \int_z^{\text{top}} dz' \int dr \, r^2 \, n(r, z'). \qquad (2-5)$$

and $n(r, z')$ is the droplet size distribution, so that $N(z') = \int^{\infty} z n(r, z') dz'$. This equation is usually re-expressed in terms of the liquid water content (LWC) of the cloud, defined by

$$LWC(z) = \frac{4\pi}{3} \rho_w \int r^3 \, n(r, z) dr, \qquad (2-6)$$

where $\rho_w = 10^6 g/m^3$ is the density of water. The effective radius is defined by

$$\frac{LWC(z)}{r_{\text{eff}}(z)} = \frac{4\pi}{3} \rho_w \int r^2 n(r, z) dz. \qquad (2-7)$$

Therefore one finally writes the optical depth in the form

$$\tau(z) = \frac{3}{2} \frac{1}{\rho_w \, r_{\text{eff}}} \int_z LWC(z') dz'. \qquad (2-8)$$

If z is below the cloud, then the total optical thickness of the cloud is

$$\tau^* = \frac{3}{2} \frac{LWP}{r_{\text{eff}} \, \rho_w}. \qquad (2-9)$$

where the liquid water path is

$$LWP = \int_{\text{bottom}}^{\text{top}} LWC(z') \, dz'. \qquad (2-10)$$

Crudely, then, $LWP = LWC \cdot t$, where t is the cloud thickness. The liquid water content varies greatly with cloud type, as shown in Table 2[4], as does the mean cloud thickness. Finally, when all of this is put together, and the average solar zenith angle of $60°$ is included, the previously inferred

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Table 2

	LWC (g/m^3)	t (km)
Cumulus	.4	2
Cumulonimbus		
Tropical	1.0	5
Trade Wind	1.5	2
Midlatitude	1.5	2.5
Polar	1.5	2
Altostratus/Altocumulus	.1	.5
Stratus/Stratocumulus	.2	1
Nimbostratus	.1	3

average cloud albedo of .45 results from an average optical thickness $\tau^* = 6$, assuming a simple geometry of plane parallel clouds. Since in fact clouds are horizontally variable, and there is some absorption, this value is actually a lower bound.

Experimentally measured cloud albedos vary greatly, as a function of type, latitude, solar zenith angle, and other parameters. But overall, they are not inconsistent with the inferred value of .45. Therefore, we may have some confidence in the value of τ^* obtained above.

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From the formula for τ^*, we see that it varies inversely with the effective droplet radius. Thus a cloud having the same LWP as another but a large number of smaller droplets, will have a larger τ^* and therefore a larger albedo.

If one asks, therefore, what changes may take place in cloud albedo (to which, as we've remarked earlier, the earth's surface temperature is extremely sensitive) due to man's activities, we must concentrate on r_{eff}. For example, an increase in r_{eff} of 10% (which corresponds to a decrease in ΔN of 30%) reduces the albedo enough so that ΔT surface $= 1.3°C$. Or, as another example, reducing N by a factor of 2 is equivalent to doubling the CO_2 concentration in the atmosphere.

A further effect of reducing droplet size, while keeping the total LWC constant, is likely to be an increase in cloud lifetime, and consequently of average cloud cover. Figure 2[5] shows that typical droplet sizes are now 5-10 μm. These do not rain out until they coalesce to form larger drops of 50 μm diameter or more. Therefore if the mean droplet size were to decrease, the time to coalesce to a size which rain will grow, thus increasing average cloud lifetime.

N and r_{eff} are largely determined by the number of CCN available. These vary widely between land and water (Figure 3)[6].

Overall, from the foregoing discussion of cloud albedo, any model purporting to predict global climate change due to greenhouse gas forcing, or anything else, needs to be able to evaluate the fractional surface areas of the earth covered by (particularly low) clouds, the liquid water content of the
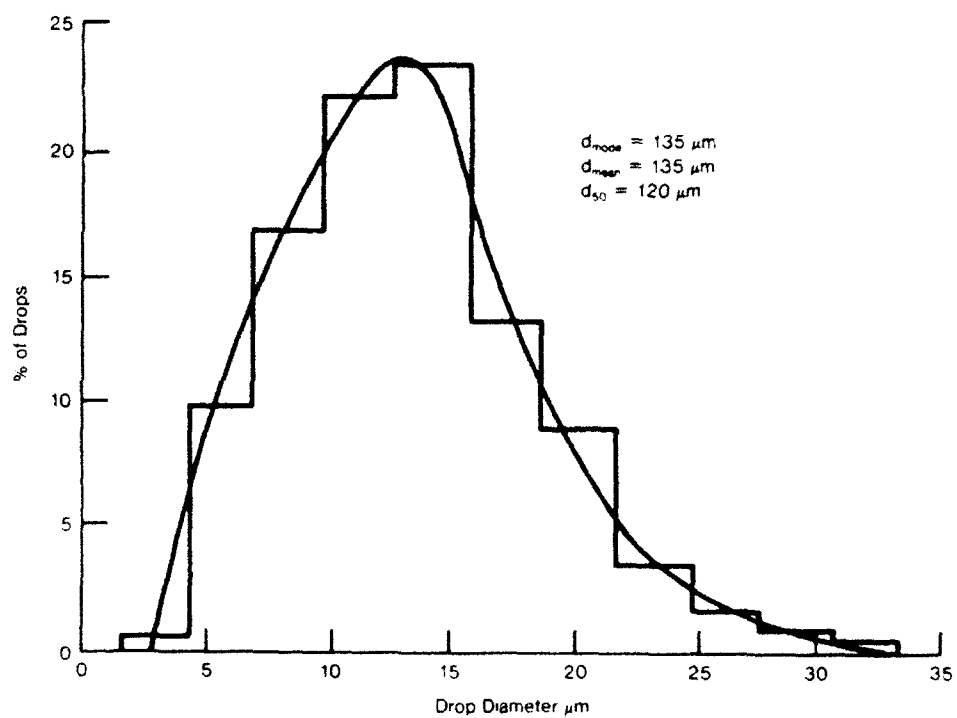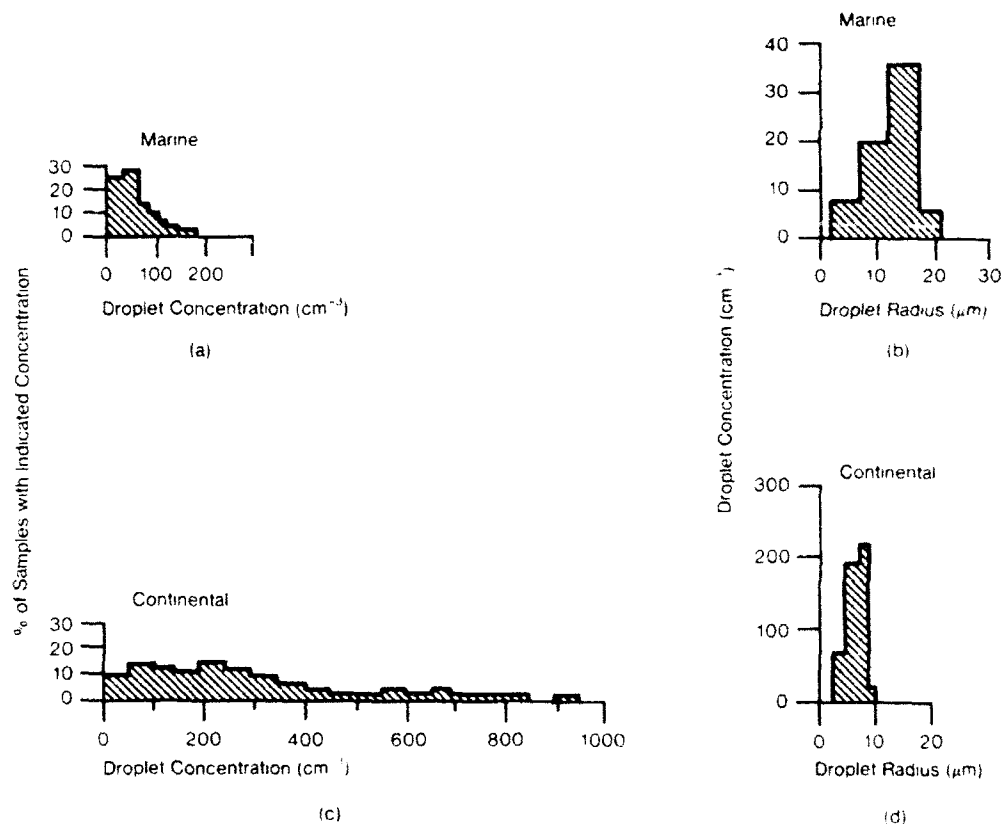
Figure 2. Average drop size spectrum for the arctic stratus clouds.

Figure 3. (a) Percentage of marine cumulus clouds with indicated droplet concentrations (b) Dropiet size distributions in a marine cumulus cloud. (c) Percentage of continental cumulus clouds with indicated droplet concentrations. (d) Droplet size distributions in a continental cumulus cloud Note change in ordinate from (b)

clouds, the droplet effective radius, the droplet number, and the number of CCN, to an extraordinarily high degree of precision; less than 5% accuracy will be required in all of these quantities.

This is an exceedingly stringent requirement on GCMs, and on computing capacity. Nothing like this precision is now available, and, indeed, many of these parameters are not even included in present models.

3 CLOUD CONDENSATION NUCLEI OVER THE OCEAN

As greenhouse gases in the atmosphere increase, the earth warms, evaporation increases, and the liquid water content of all clouds except cirrus increases. Therefore, both the albedo and the absorption of outgoing IR radiation increases. Most GCMs predict that the net effect is a positive feedback, because of the simple observation that an increase of temperature at low altitude will decrease low clouds while a decrease of temperature at high altitude will increase high clouds. Various effects, however, may reverse this.

The impact of cirrus clouds is one uncertainty. Their effect is difficult to compute, since they are composed entirely of ice crystals, which are anisotropic and whose effect on radiation is not quantitatively well understood. They are also not well studied experimentally; they are very high and also often even difficult to see visually.

It is also unlikely that fractional cloud cover will remain unchanged if evaporation increases. Generally, an increase in cloud cover will be a negative feedback. Figure 4[7] shows measured changes in average cloud cover over the ocean from 1952 and 1980, as a function of latitude, annually averaged. Could this be due to global warming?

A major uncertainty is the effect of global warming on the number of CCN, particularly oceanic CCN.

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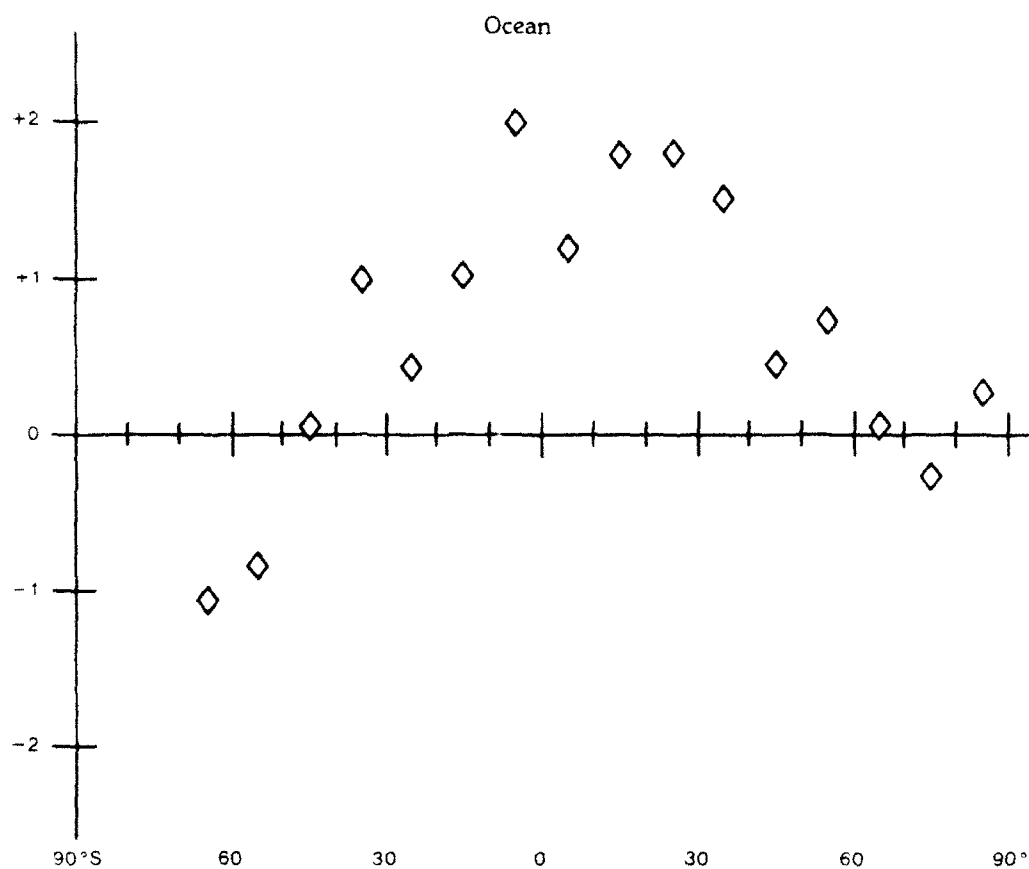
Figure 4.   Change in % • cloud-cover from 1952 to 1981 (Annual).

Marine CCN have two major components: sea salt and non-sea salt sulfate (abbreviated NSS). By volume, these are comparable. But the sea salt component is composed of much larger particles, so by number NSS completely dominates (see Figure 5)[8].

To act as a CCN, an aerosol particle must be hydrophilic and above a critical size, which is a function of the degree of supersaturation of water vapor in the cloud. Over the ocean the critical size is thought to be in the range of .05 to .14 μm, corresponding to supersaturation of 0.1% to 0.5%. (See Figure 5 again.)

Empirically, the number of CCN over the ocean is around 100 per cm^3[9] (to be compared with tens of thousands per cm^3 over polluted land areas). and since this is about the same as the droplet number density in marine clouds, it is thought that the number of CCN available is a limiting factor in the growth of marine clouds.

If we accept the idea that NSS are the dominant CCN, we must next ask how a change in global climate will affect the number of NSS particles. (We recall the apparently very strong correlation with the ice age ending 15,000 years ago shown in Figure 6.)[9]

The present concept is that NSS is produced by the oxidation of various sulfur gases coming from the surface of the sea[9]. Dimethyl sulfide ($(CH_3)_2S$, abbreviated DMS) is alleged to be the major oceanic source, and it is believed to be of biologic origin. It has a measured concentration of about 100ng/liter at the sea surface, which varies relatively little over the whole
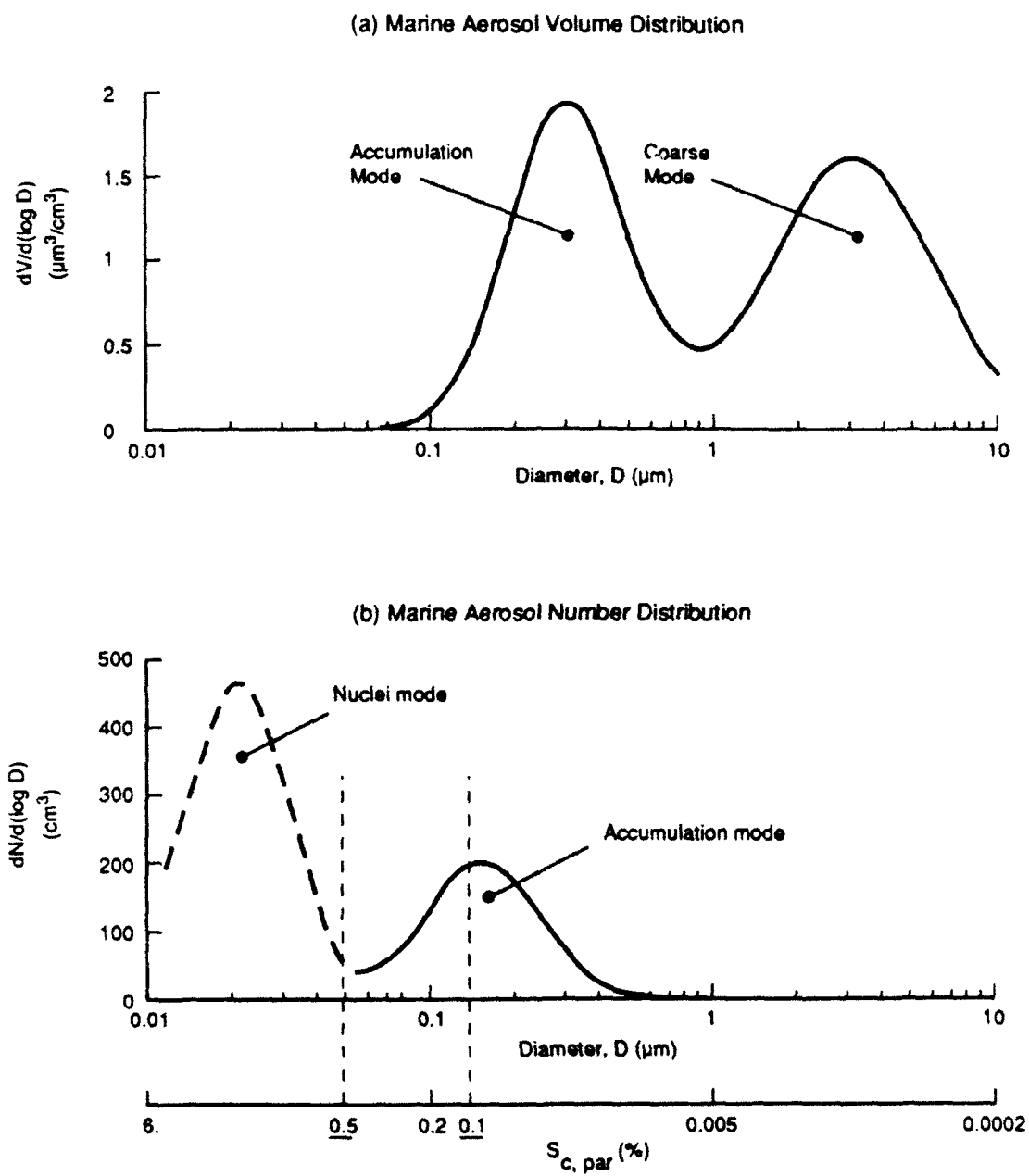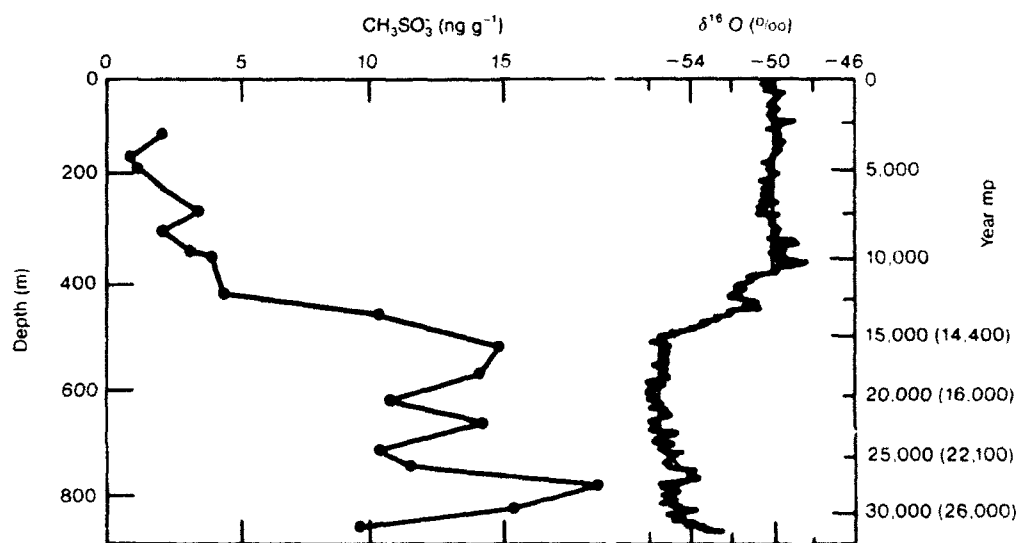
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(a) Marine Aerosol Volume Distribution

(b) Marine Aerosol Number Distribution

Figure 5.

Figure 6. MSA (CH₃SO₃) down the Dome C core; the isotope profile (δD⁰/oo) indicate the Holocene and the end of the last ice age.

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ocean surface (by a factor of no more than about 2). The entire chain of marine biochemistry leading to this concentration of DMS is very complex. involving a number of other molecules and many types of marine animals and plants. In Figure 7[9], displayed is the presently conjectured life cycle of DMS in the sea.

For our purposes it is not necessary to understand this intricate network of connections. The critical question is whether or not oceanic warming would increase or decrease DMS production, and with it, the number of CCN and hence both the albedo and cloud lifetime.

Figure 6 suggests a positive feedback; a colder environment produces an increase in sulfate CCN, which in turn increased cloud albedo which further cooled the earth. But in fact which way DMS production will change if the earth were to warm is not yet known. The effect could be either a positive or a negative feedback, and, given the extreme sensitivity of the earth's surface temperature to cloud albedo, as discussed earlier, the feedback effect could be very important.

Evidently, further research is needed on this topic, and very likely it must be incorporated in some way or other in computer models.

Finally, the existence of this potentially large feedback mechanism, so recently discovered, suggests that there are other important feedbacks which are still unnoticed. We are not yet ready for computers to hand us believable predictions about future climates, even on very gross scales.
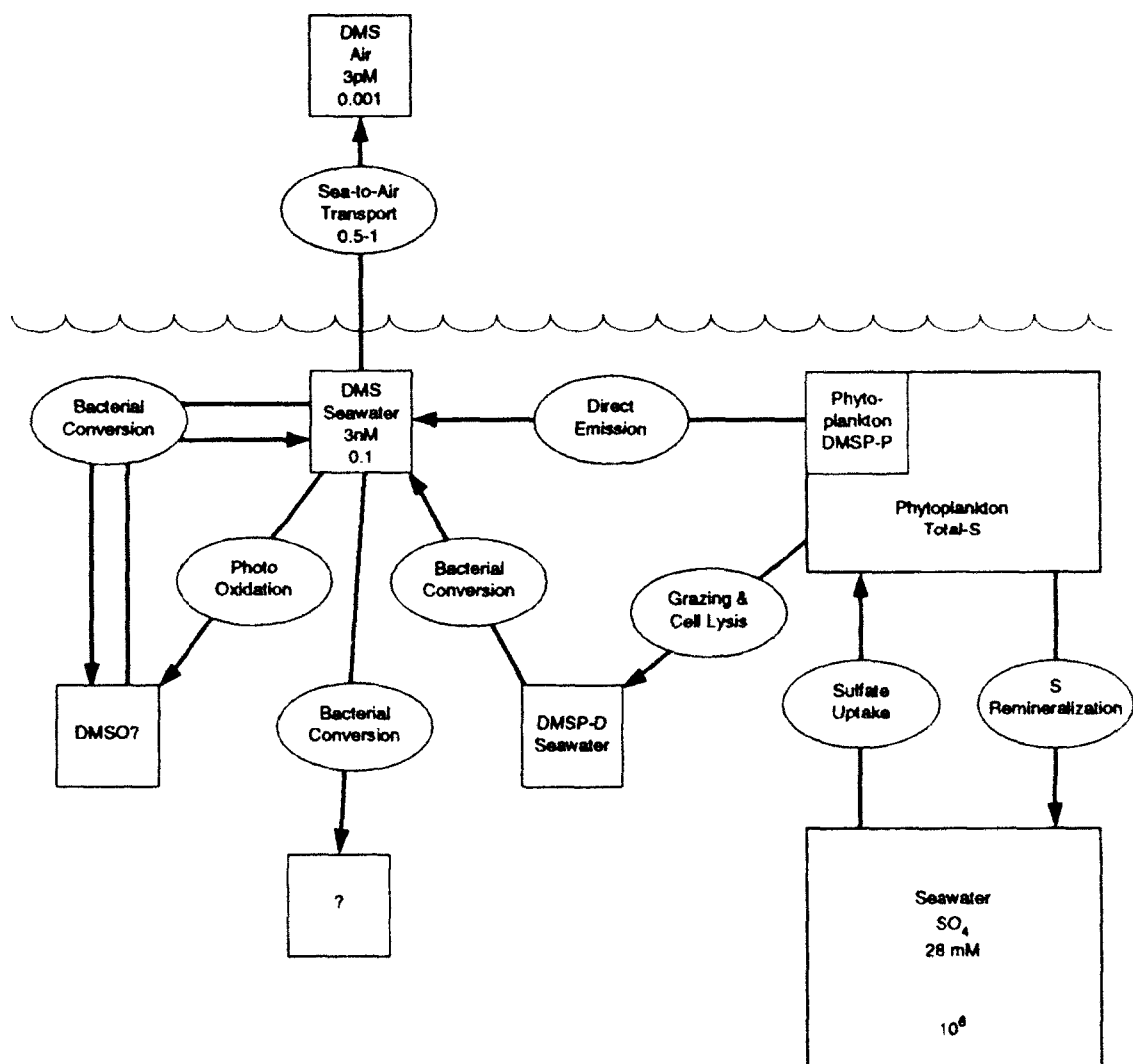
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DMS
Air
3pM
0.001

Sea-to-Air
Transport
0.5-1

Bacterial
Conversion

DMS
Seawater
3nM
0.1

Direct
Emission

Phyto-
plankton
DMSP-P

Phytoplankton
Total-S

Photo
Oxidation

Bacterial
Conversion

Grazing &
Cell Lysis

DMSO?

Bacterial
Conversion

DMSP-D
Seawater

Sulfate
Uptake

S
Remineralization

?

Seawater
SO_4
28 mM

10^6

Figure 7.

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